Search for Direct Stress Correlation Signatures
of the Critical Earthquake Model

G. Ouillon[1] and D. Sornette[1,2]

[1] Laboratoire de Physique de la Matière Condensée, CNRS UMR 6622 and Université de Nice-Sophia Antipolis, Parc Valrose, 06108 Nice, France
[2] Department of Earth and Space Sciences and Institute of Geophysics and Planetary Physics, University of California, Los Angeles, California
(e-mails: ouillon@aol.com and sornette@moho.ess.ucla.edu)

Abstract

We propose a new test of the critical earthquake model based on the hypothesis that precursory earthquakes are "actors" that create fluctuations in the stress field which exhibit an increasing correlation length as the critical large event becomes imminent. Our approach constitutes an attempt to build a more physically-based cumulative function in the spirit of but improving on the cumulative Benioff strain used in previous works documenting the phenomenon of accelerated seismicity. Using a space and time dependent visco-elastic Green function in a two-layer model of the Earth lithosphere, we compute the spatio-temporal stress fluctuations induced by every earthquake precursor and estimate, through an appropriate wavelet transform, the contribution of each event to the correlation properties of the stress field around the location of the main shock at different scales. Our physically-based definition of the cumulative stress function adding up the contribution of stress loads by all earthquakes preceding a main shock seems to be unable to reproduce an acceleration of the cumulative stress nor an increase of the stress correlation length similar to those observed previously for the cumulative Benioff strain. Either earthquakes are "witnesses" of large scale tectonic organization and/or the triggering Green function requires much more than just visco-elastic stress transfers.

1 Introduction

Numerous reports of precursory geophysical anomalies preceding earthquakes have fueled the hope for the development of forecasting or predicting tools. The suggested anomalies take many different forms and relate to many different disciplines such as seismic wave propagation, chemistry, hydrology, electro-magnetism and so on. The most straightforward approach consists in using patterns of seismicity rates to attempt to forecast future large events (see for instance [Keilis-Borok and Soloviev, 2002] and references therein).

Spatio-temporal patterns of seismicity, such as anomalous bursts of aftershocks, quiescence or accelerated seismicity, are thought to betray a state of progressive damage or of organization within the earth crust preparing the stage for a large earthquake. There is a large literature reporting that large events have been preceded by anomalous trends of seismic activity both in time and space. Some works report that seismic activity increases as an inverse power of the time to the main event (sometimes refered to as an inverse Omori law for relatively short time spans), while others document a quiescence, or even contest the existence of such anomalies at all.

There is an almost general consensus that those anomalous patterns, if any, are likely to occur within days to weeks before the mainshock and probably not at larger time scales [Jones and Molnar, 1979]. With respect to spatial structures, the precursory patterns are very often sought or observed in the immediate vicinity of the mainshock, i.e., within distances of a few rupture lengths from the epicenter. The most famous observed pattern is the so-called doughnut pattern. Thus, in any case, both temporal and spatial precursory patterns are usually thought to take place at short distances from the upcoming large event.

In the last decade, a different concept has progressively emerged according to which precursory seismic patterns may occur up to decades preceding large earthquakes and at spatial distances many times the main shock rupture length. This concept is rooted in the theory of critical phenomena (see [Sornette, 2000] for an introduction and a review adapted to a general geophysical readership) and has been documented and advocated forcefully by the russian school [Keilis-Borok, 1990; Keilis-Borok and Soloviev, 2002]. Probably the first report by Keilis-Borok and Malinovskaya [1964] of an earthquake precursor (the premonitory increase in the total area of the ruptures in the earthquake sources in a medium magnitude range) already featured very long-range correlations (over 10 seismic source lengths) and worldwide similarity. More recently, Knopoff et al. [1996] have also discovered a surprising long-range spatial dependence in the increase of medium range magnitude seismicity prior to large earthquakes in California. From a theoretical point of view, its seismological roots dates back to the branching model of [Vere-Jones, 1977]. A few years later, Allègre et al. [1982] proposed a percolation model of damage/rupture prior to an earthquake, emphasizing the multi-scale nature of rupture prior to a critical percolation point. Their model is actually nothing but a rephrasing of the real-space renormalization group approach to a percolation model performed by Reynolds et al. [1977]. Similar ideas were also explored in a hierarchical model of rupture by

Smalley et al. [1985]. Sornette and Sornette [1990] proposed an observable consequence of the critical point model of Allègre et al. [1982] with the goal of verifying the proposed scaling rules of rupture. Almost simultaneously but following apparently an independent line of thought, Voight [1988; 1989] introduced the idea of a time-to-failure analysis in the form of an empirical second order nonlinear differential equation, which for certain values of the parameters leads to a time-to-failure power law of the form of an inverse Omori law. This was used and tested later for predicting volcanic eruptions. Then, Sykes and Jaumé [1990] performed the first empirical study reporting and quantifying with a specific law an acceleration of seismicity prior to large earthquakes. They used an exponential law to describe the acceleration and did not use or discuss the concept of a critical earthquake. Bufe and Varnes [1993] reintroduced a time-to-failure power law to model the observed accelerated seismicity quantified by the so-called cumulative Benioff strain. Their justification of the power law was a mechanical model of material damage. They did not refer nor discussed the concept of a critical earthquake. Sornette and Sammis [1995] was the first work which reinterpreted the work of Bufe and Varnes [1993] and all the previous ones reporting accelerated seismicity within the model of a large earthquake viewed as a critical point in the sense of the statistical physics framework of critical phase transitions. The work of Sornette and Sammis [1995] generalized significantly [Allègre et al., 1982; Smalley et al., 1985] in that their proposed critical point theory does not rely on an irreversible damage but refers to a more general self-organization of the stress field prior to large earthquakes. In addition, using the insight of critical points in rupture phenomena, Sornette and Sammis [1995] proposed to enrich the power law description of accelerated seismicity by considering complex exponents (i.e., log-periodic corrections to scaling [Newman et al., 1995; Saleur et al., 1996; Johansen et al., 1996; 2000; Ouillon and Sornette, 2000]). This concept has been elaborated theoretically to accomodate both the possibility of critical self-organization (SOC) and the critical earthquake concept [Huang et al., 1998]. Bowman et al. [1998] gave empirical flesh to these ideas by showing that all large Californian events with magnitude larger than 6.5 are systematically preceded by a power-law acceleration of seismic activity in time during several decades, in a spatial domain about 10 to 20 times larger than the impending rupture length (i.e., of a few hundreds kilometers). The large event could thus be seen as a temporal singularity in the seismic history time-series. Such a theoretical framework implies that a large event results from the collective behaviour and accumulation of many previous smaller-sized events. Similar results were also obtained by Brehm and Braile [1998, 1999] for other earthquakes. Jaumé and Sykes [1999] have reviewed the critical point concept for large earthquakes and the data supporting it. The additional results of Ouillon and Sornette [2000] on mining-induced seismicity, and Johansen and Sornette [2000] in laboratory experiments, brought similar conclusions on systems of very different scales, in good agreement with the scale-invariant phenomenology reminiscent of systems undergoing a second-order critical phase transition. In this picture, the system is subjected to an increasing external mechanical sollicitation. As the external stress increases, micro-ruptures occur within the medium which locally redistribute stress, creating

stress fluctuations within the system. As damage accumulates, fluctuations interfere and become more and more spatially and temporally correlated, i.e., there are more and more, larger and larger domains that are significantly stressed, and thus larger and larger events can occur at smaller and smaller time intervals. This accelerating spatial smoothing of the stress field fluctuations eventually culminates in a rupture which size is of the order of the size of the system. This is the final rupture of laboratory samples, or earthquakes breaking through the entire seismo-tectonic domain to which they belong. This concept was verified in numerical experiments led by Mora et al. [2000, 2001], who showed that the correlation length of the stress field fluctuations increases significantly before a large shock occurred in a discrete numerical model. More recently, Bowman and King [2001] have shown with empirical data that, in a large domain including the impending major event, similar to the critical domain proposed in Bowman et al. [1998], the maximum size of natural earthquakes increased with time up to the main shock. If one assumes that the maximum rupture length at a given time is given by (or related to) the stress field correlation length, then this last work shows that this correlation length increases before a large rupture. Sammis and Sornette [2002] summarized the most important mechanisms creating the positive feedback at the possible origin of the power law acceleration. They also introduced and solved analytically a novel simple model based on [Bowman and King, 2001] of geometrical positive feedback in which the stress shadow cast by the last large earthquake is progressively fragmented by the increasing tectonic stress. Keilis-Borok [1990] has also used repeatedly the concept of a "critical" point, but in a broader and looser sense than the restricted meaning of the statistical physics of phase transitions (see also [Keilis-Borok and Soloviev, 2002] for a review of some of the russian research in this area). The situation is however more complicated when the strain (rather than the stress) rate is imposed; in that case, the system may not evolve towards a critical point. The unifying view point is to ask whether the dissipation of energy by the deteriorating system slows down or accelerates. The answer to that question depends on a competition between the nature of the external loading, the evolution of the deterioration within the system and how the resulting evolving mechanical characteristics of the system feedback on the external loading conditions. For a constant applied stress rate, the dissipated energy rate diverges in general in finite time leading to a critical behavior. For a constant strain rate, the answer depends on the damage law [Sornette, 1989a]. For a constant applied load, Guarino et al. [2002] find a critical behavior of the cumulative acoustic energy both for wood and fiberglass, with an exponent ≈ -0.26 which does not depend on the imposed stress and is the same as for a constant stress rate.

For the Earth crust, the situation is in between the ideal constant strain and constant stress loading states and the critical point may emerge as a mode of localization of a global input of energy to the system. The critical point approach leads to an alternative physical picture of the so-called seismic cycle. From the beginning of the cycle, small earthquakes accumulate and modify the stress field within the Earth crust, making it correlated over larger and larger scales. When this correlation length reaches the size of the local seismo-tectonic domain, a very large rupture may occur,

which, together with its early aftershocks, destroys correlations at all spatial scales. This is the end of the seismic cycle, and the beginning of a new one, leading to the next large event. As earthquakes are distributed in size according to the Gutenberg-Richter law, small to medium-sized events are negligible in the energetic balance of the tectonic system, which is dominated by the largest final event. However, they are "seismo-active" (actors) in the sense that their occurrence prepares that of the largest one. The opposite view of the seismic cycle is to consider that it is the large scale tectonic plate displacements which dominates the preparation of the largest events, which can be modelled to first order as a simple stick-slip phenomenon. In that case, all smaller-sized events would be seismo-passive (witnesses) in the sense that they would reflect only the boundary loading conditions acting on isolated faults without much correlations from one event to the other.

Notwithstanding these works, the critical earthquake concept remains a working hypothesis [Gross and Rundle, 1998]: from an empirical point of view, the reported analyses possess deficiencies and a full statistical analysis establishing the confidence level of this hypothesis remains to be performed. In this vain, Zoller et al. [2001] and Zoller and Hainzl [2001, 2002] have recently performed novel and systematic spatiotemporal tests of the critical point hypothesis for large earthquakes based on the quantification of the predictive power of both the predicted accelerating moment release and the growth of the spatial correlation length. These works give fresh support to the concept.

In order to prove (or refute) that a boundary between tectonic plates is really a critical system, the use of proxies to check the existence or absence of a build-up of cooperativity preparing a large event in terms of cumulative (Benioff) strain should ideally be replaced by a direct measure of the stress field. Indeed, one should measure the evolution of the stress field in space and time in such a region, compute its spatial correlation function, deduce the spatial correlation length, and show that it increases with time as a power-law which defines a singularity when the mainshock occurs. Unfortunately, such a procedure is far beyond our technical observational abilities. First, it is well-accepted that large earthquakes nucleate at a depth of about $10 - 15km$, so it is likely that stress field values and correlations would have to be measured at such a depth to get an unambiguous signature. Moreover, the tensorial stress field would have to be measured with a high resolution in order to show evidence of a clear increase of the correlation length. As those measurements are clearly out of reach at present, we propose here a simplified method to approach such a goal. We will then consider the 4 last largest recent events that have occurred in Southern California (Loma Prieta (1989), Landers (1992), Northridge(1994), Hector Mine(1999)), and test if such a critical scenario is likely to have taken place prior to their occurrence.

Our approach constitutes an attempt to build a more physically- or mechanically-based cumulative function in the spirit of the cumulative Benioff strain used in previous works documenting the phenomenon of accelerated seismicity.

2 General methodology

As direct stress measurements of sufficient extent for the purpose of estimating a correlation legnth are clearly out of reach, our goal is to estimate indirectly the stress distribution and its evolution with time within the crust through a numerical procedure based on instrumental seismicity.

Estimating the spatial stress history within a tectonic domain requires three different kinds of data: the first one consists in the knowledge of the far-field stress and/or strain imposed on the system. The second one consists in the accurate knowledge of the Earth's crust structure and rheology. The third one consists in the knowledge of the sources of internal stress fluctuations, which are mainly related to earthquake occurrence, whatever their size. The time evolution of the spatial structure of the stress field is thus created by the superposition of both far-field and internal contributions, coupled with the rheological response of the system (which can be quite complex). Despite its apparent simplicity, the first kind of data is still largely under debate. For example, very different scenarios are still proposed for the tectonic loading of the San Andreas fault system. Moreover, the determination of the precise boundaries of the system remain a subjet of controversy and research due to the complexity associated with the fractal hierarchical organization of tectonic blocks [Sornette and Pisarenko, 2002]. Fortunately, the critical point theory ensures that one needs only to consider the correlation function of internal fluctuations, which are the ones related to earthquakes occurrence, and not the large scale effects of the boundary conditions as long as they vary slowly on the time scale of the seismic cycle. This is why we will not further consider boundary conditions anymore here.

We shall thus use earthquake catalogs as the source of information available to qualify and quantify stress field fluctuations. Usual catalogs contain parameters such as earthquake location (longitude, latitude, depth), origin time and magnitude. For example, the CALTECH catalog that we use here is considered to be complete since 1932 for magnitudes larger than about 3.5. Unfortunately, these informations are not sufficient for quantifying the spatial stress perturbations due to a given seismic event. Two major ingredients are lacking. First, we must know the details of the rupture mechanism. This includes size (length and width), strike and dip of the fault plane, as well as the slip distribution upon it (in amplitude and direction). Those informations are usually only available for spatially and temporally restricted catalogs (but which can cover a large magnitude interval), or for more extended catalogs but only for shocks of large magnitudes (for example the Harvard catalog for shocks of magnitude larger than 5.5). As there are so few such events diluted in a very large spatial and temporal domain, it is clear that we will get in this way information on the stress field structure only at very large scales. If we consider all events in a catalog, we should be able to get insight into smaller scales (as events are much more numerous and have shorter rupture lengths), but would lack the information on the source parameters. We shall opt for the option using all the observed and complete seismicity, and will define in the next section a simplified Green function giving the spatial structure of the internal stress fluctuations due to an event of any size occurring anywhere at any

time within our system. A drastic consequence will be that this Green function will be a scalar rather than the correct tensorial structure which would be accessible if we knew the details of the rupture. Our hope is that if the critical nature of rupture is a strong property, it should be detectable even with such an approximation. In addition, the superposition effect of scalars gives in general stronger fluctuations than for higher dimensional objects such as moment tensors due to the lack of dispersion along several possible directions. The existence, if any, of an increasing correlation of the stress field should thus be detectable more easily, even if not exact quantitatively.

In order to estimate reliably the stress fluctuations and their evolution with time, we also need an accurate rheological model of the local lithosphere, including knowledge of elastic constants and relaxation times for the viscous layers. These latter ingredients can be deduced from geophysical investigations, at least on a large scale. Of course, the more accurate will be this model, the more difficult and lengthy will be the estimations of the stress field perturbations, which would necessitate the use of a finite elements or boundary elements codes. As the rheological behavior of the Earth crust and lithosphere's material can be quite complex, we shall use in the following a simplifed rheological model which captures the essential features of stress transmission and relaxation within a viscoelastic layered medium.

The methodology used in this work is the following: we first choose a recent large event (to ensure a sufficiently large catalog of possible precursor events, both in time and number), occuring at time T_0 and location P_0. We read every event in the catalog which preceeds it, and compute the spatio-temporal stress fluctuations it induces in the whole space. We also estimate, through an appropriate wavelet transform (see below), the contribution of each event to the correlation properties of the stress field around location P_0 at different scales. This will provide us with the correlation length of the stress field around P_0 and its evolution with time, up to the time of occurrence of the large event.

3   Construction of the Green function

We will first consider the stress field due to a seismic source in a 3-D elastic, infinite and isotropic medium. As catalogs do not provide us with all the parameters needed to compute accurately the exact elastic solution, we will make the following assumptions.

(i) We will consider that each source is isotropic and that the stress perturbation is positive with radial symmetry around the source.

(ii) This stress perturbation $\sigma_L(r)$ is assumed to decay from the source as:

$$\sigma_L(r) = \frac{(L/2)^3}{(L/2)^3 + r^3} \ ,$$ (1)

where L is the linear size of the source (which plays the role of the rupture length in real events), and r is the distance from the source.

The size L is determined empirically using a statistical relationship between magnitudes and rupture lengths established for strike slip faults in California [Wells and Coppersmith, 1994]:

$$\log(L) = -2.57 + 0.62 \times M_l \; , \tag{2}$$

where M_l is the local magnitude and L is expressed in kilometers. To ensure that all earthquakes are treated on the same footing, this statistical relationship is also used for the events for which the information on the rupture plane is available. Note that the computed stress $\sigma_L(r)$ is always positive and does not depend on azimut, so that it does not really define a genuine stress, but can be interpreted as a kind of influence function, with L playing the role of the size of the area in which a shock will possibly influence following events.

We now take into account that the source does not occur in a purely homogeneous elastic medium, but in a two-layers viscoelastic one. The upper layer is considered as a viscoelastic medium with relaxation time τ_1. The lower layer is also taken as a viscoelastic medium (possibly semi-infinite) with relaxation time $\tau_2 < \tau_1$. We assume that earthquakes are localized within the upper (more brittle) layer, and that the quantity of interest is the scalar stress field measured in this layer, taken constant in the vertical dimension so as to ensure that the stress field is two-dimensional within the horizontal plane. The thicknesses of the layers and the existence of free surfaces are embodied in phenomenological constants defined below. The depths of the events is taken identical and we neglect any vertical variation. This amounts to calculate the stress field at this nucleation depth.

The rupture and relaxation of the stress field in the two-layer system is modeled as follows. Once an event occurs in the upper layer, the instantaneous elastic solution for the stress field is given by expression (1). Then, both layers begin to flow by viscous relaxation. The lower layer flows faster, due to a smaller relaxation time associated with a more ductile rheology. The effect of this viscous relaxation is to progressively load the upper layer and thus creates a kind of post-seismic rebound. This loading effect computed in the upper layer is assumed to be described by a function of the type:

$$f(r,t) = \sigma_L(r)[1 - C\exp(-t/\tau_2)]\; H(t) \; , \tag{3}$$

where $\sigma_L(r)$ is the elastic isotropic solution given by (1), C is a constant which quantifies the maximum quantity of stress which is transfered in the upper layer, and which depends on the geometry of the problem. If $C = 0$, no transfer occurs. $H(t)$ is the Heavyside function which ensures that the stress fluctuation becomes non-zero once the event has occurred. Here, t is the time elapsed since the seismic event. At the same time, the stress also relaxes in the upper layer, at a rate which varies as $\exp(-t/\tau_1)$. This relaxation takes into account the usual viscous processes as well as the effect of micro-earthquakes which dissipate mechanical energy.

As both relaxations occur simultaneously, the evolution of the stress field in the upper layer is given by the sum of two contributions: (1) the direct relaxation $\sigma_L(r)\exp(-t/\tau_1)$ of the instantaneous elastic stress load in the upper layer due to the event and (2) the convolution of the time-derivative of $f(r,t)$ with the exponen-

tial relaxation function $\exp(-t/\tau_1)$ in the upper layer. This second contribution sums over all incremental stress sources $df(r,t)/dt$ per unit time in the upper layer stemming from the relaxation of the lower layer. After some algebra, we get the stress perturbation induced by an earthquake under the form

$$\sigma(r,t) = \frac{(L/2)^3}{(L/2)^3 + r^3} \left[\exp(-t/\tau_1) + B\frac{\tau_1}{\tau_1 - \tau_2} \left(\exp(-t/\tau_1) - \exp(-t/\tau_2) \right) \right] , \quad (4)$$

where r and t are respectively the horizontal distance from the source and the time since the occurrence of the earthquake. The constant B represents the relative contribution to the stress field in the upper layer due to the delayed loading by the slow viscous relaxation of the lower layer that has been loaded by the instantaneous elastic stress transfer at the time of the earthquake compared with the direct relaxation of the elastic stress created directly in the upper layer. The numerical value of B is difficult to ascertain as it depends strongly on the geometry of the layers as well as on their rheological constrast. We expect both contributions to be of the same order of magnitudes and, in the following, we shall take $B = 1$.

The Green function defined here is a rough approximation of what really takes place within the crust and the lithosphere, but it nonetheless captures qualitatively the overall evolution of the stress field. One could raise the criticism that it does not feature any azimutal dependence of the stress field perturbation but, as we already stated, this is done in view of the absence of detailed information on the source mechanisms of the events. On the other hand, as stated above, the use of an isotropic stress field is expected to lead to an overestimation of the correlation length, that is, to an amplification of the signal we are searching for. While we cannot provide a rigorous proof of this statement, it is based on the analogy between percolation and Anderson localization [Souillard, 1987, Sornette, 1989b,c]: the first phenomenon describes the transition of a system from conducting to isolating by the effect of the addition of positive-only contributions; The second phenomenon refers to the transition from conducting to isolating when taking into account the "interferences" between the positive, negative and more generally phase-dependent amplitudes of the electronic quantum wave functions. In this later case, the transition still exists but is much harder to obtain and to observe. In the future, it may nevertheless be interesting to check this point and test a generalization of the present model in which a random source orientation is chosen for each event and the angular dependence of the associated double-couple stress is taken into account.

The Green function we propose also assumes a complete decoupling between space and time, so that viscous relaxation does not exhibit any diffusive pattern. Indeed, such a diffusion would imply an increase of the size of the influence area with time. As the amplitude of the stress signal decreases exponentially with time, we believe that this mechanism is not crucial (because too slow and too weak in amplitude) in order to obtain and measure an increase of the stress field correlation length, if any. Another assumption of our rheological model is that the viscoelastic component is linear, allowing to clearly define relaxation times. This ingredient allows us to define a simple and convenient computation procedure to estimate a correlation length, as

discussed in the next section.

The simplified Green function $\sigma(r,t)$ given by (4) has several interesting properties catching the overall physics of the stress evolution in the upper layer after an event. The elastic prefactor $\sigma_L(r)$ given by (1) implies that the stress perturbation is initially of order unity within a circle of radius $L/2$, and sharply decreases as r^{-3} outside this circle. Note that the maximum amplitude of the stress perturbation is independent of the size L, as the stress drop is thought to be constant, whatever the size of an event. At any point in the upper layer, the stress will first be given by the elastic solution. As $\tau_1 > \tau_2$, the stress at any point in the upper layer will first increase due to the relaxation of the lower layer, reach a maximum, and then decrease with time as the upper layer is relaxing too, but with a longer relaxation time.

Figure 1 shows such a scenario with $\tau_1 = 10$ years and $\tau_2 = 1$ year. The maximum amplitude depends on the distance between the event and the point where this stress is measured (as well as on B).

If we now superimpose the contributions of all successive earthquakes in a catalog, the stress history at any given point will be a succession and/or superposition of such fast growing and slowly decaying stress pulses. We thus construct the cumulative stress function $\Sigma(t)$ defined as

$$\Sigma(t) = \sum_i \sigma(r_i, t_i) \; , \tag{5}$$

where $\sigma(r_i, t_i)$ is given by (4) and r_i and t_i are the distance and the time of event i to the main shock. For example, Figure 2a shows the stress history measured at the location of the Landers epicenter due to the succession of all previous events in the catalog, assuming $\tau_1 = 1$ and $\tau_2 = 6$ months. Figure 2b shows the same computation for $\tau_1 = 10$ years, while Figure 2c assumes $\tau_1 = 100$ years. Increasing τ_1 widens the stress pulses, which lead them to overlap and produces a more continuous stress history.

The constructions of $\Sigma(t)$ shown in Figure 2a-c are analogous to the cumulative Benioff strain studied in [Bufe and Varnes, 1993; Sornette and Sammis, 1995; Bowman et al., 1998; Brehm and Braile, 1998; 1999; Jaumé and Sykes, 1999; Ouillon and Sornette, 2000], and are an attempt to improve upon them as we now explain. They are analogous because they can be seen as similar to the sums of the type

$$M_q(t) = \sum_{i|t_i<t} [M_0(i)]^q \; , \tag{6}$$

where $M_q(t)$ is a moment generating function of order q, t_i and $M_0(i)$ are the time and seismic moments of the i-th earthquake and q is an exponent usually taken between 0 and 1. The cumulative Benioff strain is obtained as $M_{q=1/2}(t)$ where the sum is performed over all events above a magnitude cut-off in a pre-defined spatial domain. Taking $q = 1$ corresponds to summing the seismic moments, while taking $q = 0$ amounts to simply constructing the cumulative number of earthquakes. The constructions shown in Figure 2a-c can be seen as equivalent to $M_{q=0}(t)$ when the two following limits hold: (1) all earthquakes in the catalog are so close to each

other that they are all within a distance less than their rupture length from the point where the stress is calculated (in this case, the elastic stress perturbation brought by each event is equal to the constant stress drop); (2) the time difference between the occurrence of each event and the main shock is significantly less than τ_2 such that the time-dependence in (4) can be neglected.

A significant advantage in our construction of the cumulative stress function $\Sigma(t)$ defined by (5) compared with the cumulative Benioff strain resides in the fact that we do not need to specify in advance a spatial domain, a delicate and not-fully resolved issue in the construction of cumulative Benioff strain functions. The definition of the relevant spatial domain is automatically taken into account by the spatial dependence of the Green function.

Two ingredients are going to modify the observed acceleration of the Benioff strain when studying the cumulative stress function $\Sigma(t)$ defined by (5). The first one is that each event contributes a maximum stress perturbation equal to the stress drop. In contrast, large events contribute significantly more in the cumulative Benioff strain as the square-root of their seismic moment and independently of their distance. There is however a size effect in our calculation of $\Sigma(t)$ that reveals itself at large distances $r_i \gg L_i$, stemming from the magnitude dependence of the range L_i of the stress perturbation. According to (2) and using the standard relationship between magnitude M_L and seismic moment M_0, $M_L = (2/3)[\log M_0 - 9]$, we obtain $L_i \sim [M_0(i)]^{0.4}$ and thus $\sigma(r_i, t_i) \sim L_i^3 \sim [M_0(i)]^{1.2}$ for $r_i \gg L_i$. This size effect has however an almost negligible contribution in generating an acceleration because the stress field becomes small at large distances. The second ingredient limiting the acceleration of the cumulative stress function $\Sigma(t)$ defined by (5) is the relaxation in time which is responsable for the decay observed in Figures 2a-c. The longer τ_1 is, the smaller is the amplitude of this decay, until $\Sigma(t)$ is replaced by a staircase in the limit $\tau_1 \to +\infty$. The largest values of τ_1 that we have explored are significantly larger than the total duration of the catalog and larger values will not change our results quantitatively.

Another important issue is the contribution of the small events not taken into account in the sum (5). Indeed, the typical area $S(L)$ over which the stress redistribution after an event is significant is of the order of the square $S(L) \propto L^2$ of the size L of the rupture. If the earthquake seismic moments M are distributed according to a density Pareto power law $\propto 1/M^{1+\beta}$ with $\beta \approx 2/3$ (which is nothing but the Gutenberg-Richter law for magnitudes translated into moments), using the fact that $M \propto L^3$, the density distribution of the areas $S(L)$ is also a power law $\propto 1/S^{1+(3/2)\beta}$ with an exponent $(3/2)\beta \approx 1$. Thus, the contribution of each class of earthquake magnitudes is an invariant: small earthquakes contribute as much as large earthquakes to the sum (5). Therefore, it seems a priori very dangerous to ignore them in our sum (5) which attempts to detect a build-up of correlation. However, if we assume that the physics of self-organization of the crust prior to a critical point is self-similar, the critical behavior should be observable at all the different scales and neglecting the contribution of small events should not lead to a destruction of the signal nor to a modification of its relative variations, only to a change in its absolute

amplitude.

To sum up, our physically-based definition of the cumulative stress function adding up the contribution of stress loads by all earthquakes preceding a main shock seems to be unable to reproduce an acceleration similar to those observed previously for the cumulative Benioff strain. This is due to the fact that, conditioned on the hypothesis of a magnitude-independent stress drop and using standard elasticity, the impact of the largest events is not significantly larger than those of smaller events. In view of this failure, we now attempt another hopefully more robust characterization of the critical point model.

4   Analysis of the structure of the stress field

Our objective is to determine the correlation length of the computed stress field in the neighboorood of 4 large shocks in California as a function of the time before their occurrence. In this goal, we are going to analyze the structure of the stress field around each main shock epicenter to check whether the stress fluctuations are increasing or decreasing in size around each main shock epicenter. In order to extract a robust estimation of the correlation length of the stress field reconstructed from a limited number of events, we investigate what spatial scales or wavelengths are developping around each main shock epicenter, that is, what is the characteristic scale of the roughness of the computed stress field.

An efficient way to achieve such a goal is to perform a $2D$ wavelet transform of the stress field, which acts as a microscope allowing us to focus on separate scales. As we are interested only in the spatial structure surrounding the upcoming mainshock (defined as point P_0), we compute wavelet coefficients centered at location P_0. We consider the following wavelet:

$$\frac{1}{a}(2 - \frac{r_p^2}{a^2}) \exp(-\frac{r_p^2}{2a^2}) \tag{7}$$

centered at point P_0. This "Mexican hat" wavelet is the second-order derivative of the Gaussian function. By construction, it eliminates signals of constant amplitudes or of constant gradient at scale a or larger. It is thus well indicated to isolate fluctuations at various chosen scales. r_p is the distance to point P_0, and a is the analyzing scale (the larger a, the larger the width of the wavelet). Note that working with a scale a means that the corresponding structures have in fact a size $2.2a$ [Ouillon, 1995].

For each time in the stress field history, the wavelet transform is obtained by convolution of this function with the computed spatial stress field, for different values of a. If the resulting wavelet coefficient is close to 0, this means that the stress field is uniform or varies linearly around P_0, at scale a. If the coefficient is strongly negative, this means that P_0 is at or near a local stress minimum, at scale a. If it is strongly positive, this means that P_0 is at or near a stress maximum at scale a, indicating that the stress is both locally high and correlated at that scale. This is exactly the property that we want to check.

Our analyzing procedure is thus the following: we consider the first event in the catalog. We compute the stress field fluctuation due to this event at any time and any location through equation (4). The wavelet transform provides the contribution of this event at any time to the total wavelet coefficient at any scale a at location P_0. Summing all contributions of successive events (as the rheology we chosen is linear) up to the major mainshock provides us with the complete evolution of the scale content of our computed stress field around P_0. From the wavelet coefficient of the cumulative stress field as a function of scale at a fixed time t, we extract the corresponding correlation length $\rho(t)$ as the scale corresponding to the maximum coefficient, multiplied by 2.2. If the critical point hypothesis is correct, $\rho(t)$ should behave as

$$\rho(t) = A + C(T_0 - t)^{-\nu} ,\qquad(8)$$

where ν is a positive critical exponent. Note that, due the very small rupture size L for small earthquakes, and as the scale a varies from 1 to $100km$, it would be necessary to grid a very large domain (of few hundreds kilometers large) with a very small mesh size (of order few tens of meters). This would make computations and data storing practically untractable. This is why we have defined a procedure which computes data only on very small subgrids whose size (and mesh size) depends on the wavelet scale and on the event size. This procedure is made possible because we compute wavelet coefficients at several scales but only at a single location, namely the position of the upcoming large event. Indeed, we do not store the stress history for all locations, but only at the position P_0 of the epicenter of the target main shock.

5   Results

We have analyzed the evolution of the stress field before 4 large Southern Californian shocks: Loma Prieta (1989), Landers (1992), Northridge (1994) and Hector Mine (1999). We restricted our analysis to those 4 recent events as this ensures that our computed stress field history is the longest possible for this area, and is not subjected to finite size effects (as these 4 events are located at the end of the catalogue). The Caltech catalog we used is thought to be complete since 1932 for events of magnitude larger than 3.5. Computation of the stress field before each of the selected large events included all events of magnitude larger than 4 since 1932.

Three parameters dictate the properties of the Green function of a seismic event in our computations, namely the relaxation time scales τ_1 and τ_2, and the stress amplification factor B. We made several computations, varying those 3 parameters. We checked that the less influential parameter is B. Another parameter which has a rather low influence on the results is τ_2, the relaxation time of the lower, more ductile medium. The most influential parameter is τ_1, the relaxation time of the upper layer. If τ_1 is too small, then all events appear as very well individualized temporal stress pulses decaying very fast before the next event takes place. As a consequence, the dominating space scale is never defined, except at the time of occurrence of each event, where it is of the order of the distance between this event and P_0. The optimal

space scale thus varies very wildly with time.

When increasing τ_1, stress pulses gradually overlap in time. Finally, when τ_1 is infinite, stress pulses becomes steps without any relaxation. Increasing τ_1 leads to a less erratic behaviour of the optimal spatial scale length obtained from our wavelet analysis. We will here consider a Green function with relaxation times $\tau_1 = 100$ years and $\tau_2 = 0.5$ year. The scalar stress history computed at the location of the Landers shock is shown on Figure 2c. It globally increases with time (as all previous events stress perturbations are positive by definition) but does not exhibit any acceleration. Note that stress steps (due to neighboring events) are followed by a smooth decay, due to the very slow relaxation associated with the high τ_1 value. The time step for the computation of each successive point of the cumulative stress is 6 months. We stress that the procedure we use provides results independent of the time step, thanks to our linear rheology.

Figure 3 shows the wavelet coefficients for the cumulative stress function constructed for the Landers 1992 earthquake as a function of scale at various times. Time increases from the bottom to the top (the very upper curve has been computed just before the Landers shock). The curves with the lowest amplitudes, corresponding to the early years, are flat as the number of shocks is low, so that the stress field is almost 0 everywhere, and no specific structure emerges as too few events have been included in the computation. Later, the amplitude of the profile increases in amplitude (either positively or negatively), but it is worth noting that its shape is almost constant. As time increases, the amplitude of the stress field varies, but its structure remains constant, at least at point P_0. For example, for wavelet scales lower than $10km$ (true size lower than $22km$), the "future" Landers epicenter is found to be located in a local stress deficit. The local correlation length of the stress field, given by the maximum of the wavelet coefficient, occurs for a constant scale of about $25km$ (true size of about 55 km). We note that this maximum occurs at the same scale for all times. Figure 4 shows the evolution with time of the correlation length. It first fluctuates widely, as there are too few events to compute a representative stress fluctuations field, but then enters a very stable phase with no noticeable variation with time. We thus show no increase or decrease of this local correlation length, which confirms the fact that the local structure of the computed stress field does not exhibit any major change when approaching failure around P_0.

Figures 5 to 7 show the results of the same computations before the Loma Prieta event. The correlation length is found constant from 1958 to 1987, with a value of about $77km$ (wavelet scale of $35km$).

Figures 8 to 10 show the results of the same computations before the Northridge event. The correlation length is found constant from 1972 to 1994, with a value of about $66km$.

Figures 11 to 13 show the results of the same computations before the Hector Mine 1999 event. Once again, no clear increase of the correlation length occurs before the large event.

We also performed the same tests considering only catalog events of magnitude larger than 5. We obtain exactly the same results, except that the wavelet profiles of

Figures $3, 6, 9$ and 12 are found to be dilated along the scale axis. This just reflects the fact fewer events are taken into account in the computations, and are thus more diluted in space. We also performed tests using a larger distance of influence of each given event quantified by equation (1)) by doubling the rupture size $L \rightarrow 2L$. The results are qualitatively the same.

6   Interpretation and discussion

Using simplified models of earthquake elastic stress transfer and of the lithosphere rheology, we have attempted to model the stress field evolution from 1932 up to the occurrence time of recent large Southern Californian events. This allowed us to analyze the time evolution of our simplified cumulative stress field at the locus of large impending shocks before their occurrence, and to determine the spatial correlation length of this local stress field. Using a variety of rheological models did not allow us to find evidence of a strong increase (nor any other peculiar variation) of both the cumulative stress field and of the correlation length before any of the 4 major events studied here. These negative results would not change by replacing the simple exponential decays by power laws of the form of the Omori law for aftershocks, since taking an infinite range correlation $\tau_1 \rightarrow +\infty$ does not change our results.

We have observed that all large events occured in a local minimum of the computed stress field at (true) scales less than $20 - 25km$, and that this minimum becomes more and more pronounced with time. A magnitude 7 event has an average rupture length of about $70km$. As we have stressed before, such an event certainly nucleates in a zone where the stress field is correlated on long wavelengths. The final length of the rupture will stem from the interplay between this initial static stress field structure and details of rupture dynamics (inertial effects coupled with the specific geometry of the rupture plane). We can guess that the final extent of the rupture will be larger than the initial correlation length of the stress field. This is why we could expect that this correlation length before each of the 4 major events should have been of the order of a few tens of kilometers. It is thus puzzling to observe that the wavelet coefficients at scales of 10 to $20km$ are becoming more and more negative with time. This observation is perhaps due to the naive shape of the Green function we considered, which is positive everywhere. However, we believe that if this assumption certainly affects the value of the computed stress field, it should certainly lead us to an overestimation of the correlation length, as more space is filled with positive stress. We are thus forced to conclude that there is neither a strong stress field and nor large stress correlation at the scale of a few kilometers scale. It thus seems that the mechanism of stress transfer due to the occurrence of successive smaller-sized events is not a direct ingredient in building long correlations in the cumulative stress field, which are necessary for the propagation of large future events according to the critical point model.

These results are in contradiction with those reported in the literature [Bufe and Varnes, 1993; Sornette and Sammis, 1995; Bowman et al., 1998; Brehm and Braile,

1998; 1999; Jaumé and Sykes, 1999; Ouillon and Sornette, 2000] based on the cumulative Benioff strain, who showed that large-scale spatial and temporal correlations characterize seismicity before a large event in the same area.

Our results may be reconciled with those previous studies if we acknowledge that medium-sized events are not seismo active (they are not "actors"). In other words, the temporal singularities defined in [Bowman et al., 1998] for instance stem rather from the large scale geometry of the boundary loading conditions and correlations not directly mediated by the stress field (that were not taken into account in the present work) than from strong interaction between seismic events mediated by the stress field. In this spirit, King and Bowman [2001] and Sammis and Sornette [2002] have developed a model in which the main mode of loading of a previously ruptured major fault occurs by localized viscous flow beneath this fault. The consequence is that the extent of the stress shadow due to the previous mainshock decreases with time, so that seismicity migrates back to the mainshock epicenter in an accelerating manner, the temporal singularity coinciding with a new mainshock on the fault. However, such a model implies that seismicity migrates towards P_0, which cannot reasonably be inferred from our computations either (Figure $3, 6, 9$ and 12). If this was the case, the wavelet coefficients should be negative at P_0, and the width of the domain around P_0 where coefficients are negative should decrease with time. This suggests that the loading mechanism proposed by King and Bowman [2001] and Sammis and Sornette [2002] does not explain the data, but that another loading mechanism may explain the temporal singularity coinciding with large events.

Another solution to explain the discrepancy between the large scale correlations observed in seismic catalogs [Bufe and Varnes, 1993; Sornette and Sammis, 1995; Bowman et al., 1998; Brehm and Braile, 1998; 1999; Jaumé and Sykes, 1999; Ouillon and Sornette, 2000] and our results is to argue that our geometrical/rheological model of the lithosphere is incorrect, which makes our Green function imperfect. The Green function we have considered is representative of a linear viscoelatic layered medium, and we checked that our results are not strongly dependent on its various parameters. One possibility is that, if the observed absence of correlations is due to our choice of the Green function, then the true Green function must be of a fundamentally different nature. The Earth's crust is a very complex medium, composed of blocks of various sizes separated by fractures or fault zones, subjected to a confining pressure and temperature increasing with depth. We would be indeed very lucky if such a medium behaved as a perfect linear medium. Indeed, crustal rheology must be of nonlinear nature, even in its most superficial "elastic" part. Some evidence of a nonlinear response associated with the anisotropic response of a cracked medium under compression compared to tension has been reported in [Peltzer et al., 1999]. Extending this argument, if, for example, the crust behaves as a granular material, then we must expect that tectonic forces propagate over longer distances within much narrower channels than those predicted by standard elastic models. This singular property is due to the hyperbolic nature of stress propagation differential equations in granular media [Bouchaud et al., 1995; 2001], whereas those equations are of ellipitical nature in standard elasto-plastic media. The real rheology

of the Earth's crust is probably somewhere between that of a granular material and standard (possibly nonlinear) visco-elastico-plasticity. It thus seems important to better understand crustal rheology (and its associated Green function), in order to check the changes it would imply in the various brittle crustal modes of deformation and in the way earthquakes "speak to each other." In this spirit, phenomenological models of earthquake interaction and triggering are quite successful in capturing most of the phenomology of seismic catalogs [Helmstetter and Sornette, 2002; Helmstetter et al., 2002]. It remains to derive the triggering Green function from physically-based mechanisms, which seem to require much more than just visco-elastic stress transfers.

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Figure 1: Evolution with time of the time-dependent part of the normalized stress field showing the loading phase induced by the relaxing lower layer and the large time relaxation phase in the upper layer. The parameters are $\tau_1 = 10$ years, $\tau_2 = 1$ year and $B = 1$.

Figure 2: Cumulative stress function as a function of time at the location of the Landers epicenter calculated by summing the contributions $\sigma(r_i, t_i)$ given by (4) of the Green functions generated by all previous events i, that occurred at times t_i prior to the Landers earthquake taken at the origin of time and at distances r_i from the Landers epicenter. (a) $\tau_1 = 1$ year and $\tau_2 = 6$ months; (b) $\tau_1 = 10$ years and $\tau_2 = 6$ months; (c) Same as Figure 2a with $\tau_1 = 100$ years and $\tau_2 = 6$ months.

Figure 3: Wavelet coefficients for the cumulative stress function constructed for the Landers 1992 earthquake as a function of scale a at various times. Time increases from the bottom to the top (the very upper curve has been computed just before the Landers shock).

Figure 4: Correlation length estimated at the Landers epicenter of the cumulative stress function for the Landers earthquake as a function of time.

Figure 5: Same as Figure 2c for the Loma Prieta 1989 earthquake.

Figure 6: Same as Figure 3 for the Loma Prieta 1989 earthquake.

Figure 7: Same as Figure 4 for the Loma Prieta 1989 earthquake. The correlation length is found constant from 1958 to 1987, with a value of about $77km$ (wavelet scale of $35km$).

Figure 8: Same as Figure 2c for the Northridge 1994 earthquake.

Figure 9: Same as Figure 3 for the Northridge 1994 earthquake.

Figure 10: Same as Figure 4 for the Northridge 1994 earthquake. The correlation length is found constant from 1972 to 1994, with a value of about $66km$.

Figure 11: Same as Figure 2c for the Hector Mine 1999 earthquake.

Figure 12: Same as Figure 3 for the Hector Mine 1999 earthquake.

Figure 13: Same as Figure 4 for the Hector Mine 1999 earthquake.

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